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Numerical Modeling of a Case of Nocturnal Radiation Reversal

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ABSTRACT—A one-dimensional boundary-layer model is used to simulate a rare occurrence of nocturnal radiation reversal, with measured downward radiation and horizontal pressure and temperature gradients regarded as given. The time sequence of surface temperature, wind, and evaporation is well simulated except for a period of calm observed at the radiation station but not elsewhere.

1. INTRODUCTION

How best to treat the effects of turbulence in dynamic numerical simulation experiments is still being debated (e.g., Zilitinkevich 1970, pp. 190–241). Probably the least explored part of the question is the nocturnal boundary layer over land. Although this layer is of less importance in large-scale modeling, it does present difficult problems in the dynamics of stratified flow. The author (Clarke 1970) has suggested procedures for use in modeling based partly on the work of Webb (1970), and opportunities are sought to test the efficacy of these procedures. The work described here has this as its primary objective, the view being taken that a successful model should be able to simulate unusual events such as those associated with radiation reversal.

2. DESCRIPTION OF THE REVERSAL OCCURING ON OCT. 4–5, 1970

This date is one of the two occasions of reversal with strong gesotrophic north wind and middle cloud referred to in a companion paper by the author (Clarke 1972). This case of radiation reversal was selected for simulation using the model described in the following section.

The weather sequence was extraordinary in several respects. "Red rain" in Melbourne, Australia, on the evening of October 3 was associated with transport of dust aloft from the Simpson Desert, some 1300 km to the north-northwest, over a shallow layer of cooler air with a trajectory from the southwest. Early on October 4, the cooler air was swept away, and warm surface northerlies now almost completely free of dust became established. Surface temperatures rose rapidly to 29°C, the highest October temperature in Melbourne for 40 yr and to 35°C at a station some 500 km inland to the north-northwest. Surface north wind speeds in Melbourne remained generally light to moderate until 2200 LST (but a seabreeze from the southwest prevailed at Aspendale, Australia, until 1500 LST). The north wind then rose steadily to 11 m/s with gusts over 20 m/s by 0300 LST on October 5 (35 m/s at 900 mb according to a wind sounding at Laverton, Australia, 37 km west-northwest of Aspendale). A cold front arriving at Aspendale at 0600 LST on the 5th brought an end to the strong surface winds and high temperatures (hourly average value of 28.6°C at 0130 LST). The hourly periods centered on 2230, 2330, and 0030 LST averaged marked positive net radiation, while the lysimeter-measured evaporation was quite high during most of the night. An extensive sheet of clouds with base at 3000 m (estimated temperature 6°C), as measured at the airport 39 km north-northwest of Aspendale, evidently passed over during this time. Visibility at the airport remained very good, indicating little if any dust in the air.

3. THE NUMERICAL MODEL

A model, based on the recommendations of Clarke (1970, method 1) has been developed for the purpose of representing as accurately as possible boundary layer effects in dynamic simulation experiments. The model has 22 levels in the vertical with a "sigma coordinate" (Phillips 1957) whose lowest level is at $\sigma=0.9984$ (≈ 13 m), nine levels in the lowest kilometer, five levels in the next, and the highest level at $\sigma=0.0931$ (16 km).

The model computes forward in time (time step 15 s) from initial conditions (surface pressure, the fields of temperature, humidity, wind) and from the values of these variables at the lowest model level, it derives fluxes of heat, water vapor, and momentum at the surface. For this purpose, a knowledge of the downward radiative flux at the surface is required. Four equations are solved, expressing the surface energy balance and flux-gradient relationships as follows:

$$R\downarrow -\sigma T_s^4=\mathcal{L}E+H+G, \tag{1}$$

$$\frac{u}{u_*}=F_m\frac{z}{L}-F_m\frac{z_0}{L}, \tag{2}$$

$$\frac{\delta\theta}{T_*}=F_H\frac{z}{L}-F_H\frac{z_0}{L}, \tag{3}$$

and

$$\frac{\delta q}{q_*}=F_W \frac{z}{L}-F_W \frac{z_0}{L} \tag{4}$$

where $R\downarrow$ is downward radiative flux, σ is Stefan's constant, T_s is surface temperature, $\mathcal{L}$ is latent heat of vaporization, E is water vapor flux, H is sensible heat flux, G is heat flux into the ground, u_* is friction velocity, z is height, L is the Obukhov length, z_0 is roughness length, u is wind speed at the lowest model level, $\delta\theta$ is the increment in potential temperature from the surface to the lowest model level, δq is the increment from the saturation value of specific humidity at temperature T_s to the specific humidity at the lowest model level, the Fs are treated as known functions, $T_*=-H/(\rho c_p u_*)$, and $q_*=-E_{\text{pot}}/(\rho u_*)$. E_{pot} is potential evaporation, c_p is specific heat capacity at constant pressure, and ρ is density.

Four levels in the soil ranging from 1 to 80 cm are used to compute heat fluxes and temperatures in the soil. No attempt is made to simulate the movement of water substance in the soil as has been done by Sasamori (1970) using the formulation of Phillips (1957). Instead, $D=E/E_{\text{pot}}$, the ratio of actual to potential evaporation, is regarded as a quantity determined by soil moisture content, and fixed values are used for thermal conductivity and diffusivity appropriate to the quality and dampness of the soil (DeVries 1963). Two experiments were made, one of which assumed that $D=0.2$ and the other that D is a sinusoidally varying quantity ranging from 0.1 at midday to 0.8 at midnight. The differences were small, and the results quoted are those for the former experiment.

The boundary layer above the lowest model level presents difficulties to the modeler because of our lack of understanding of turbulent processes at higher levels, especially in stable conditions. In the present instance, a formulation was used that has so far been found to give the best fit to the Wangara data (Clarke et al. 1971), averaged according to hour of day. Results of these experiments will be reported elsewhere.

In the present formulation, the mixing length, l, is assumed to vary with height in the following manner:

1. When the Obukhov length, L, is >0; that is, for stable conditions, $l=kz(1+\alpha pz/L)^{-1}$ where $p=1$ when $z/L\leq 1$, and $p=(z/L)^{-1}$ when $z/L>1$; $\alpha=5$, von Kármán's constant $k=0.4$, and z is height. When l would be otherwise greater than 30 m, $l=30$m.

2. For unstable conditions ($L<0$), the profiles are searched to find the height, $z=z_i$, at which the gradient Richardson number is unity. Then $l=kz(1-16z/L)^{+0.25}$ up to $z=0.3\ z_i$, and thereafter goes linearly to 30 m at z_i, remaining at that value at higher levels.

For present purposes, the model is stripped to one dimension, no radiation is computed, and convection is simulated by the "dry convective adjustment" of Manabe et al. (1965). Downward radiation, $R\downarrow$, as measured at Aspendale, was allowed to vary in time. Horizontal gradients of wind and temperature (used for computing horizontal advection of momentum and heat) and of surface pressure were estimated from available 3-hourly surface synoptic charts and fed to the model as variables in time, linearly interpolated. For initial conditions, air temperature, humidity, and wind profiles measured at Laverton at 0900 LST on October 4, and Aspendale soil temperatures at the same time, were used.

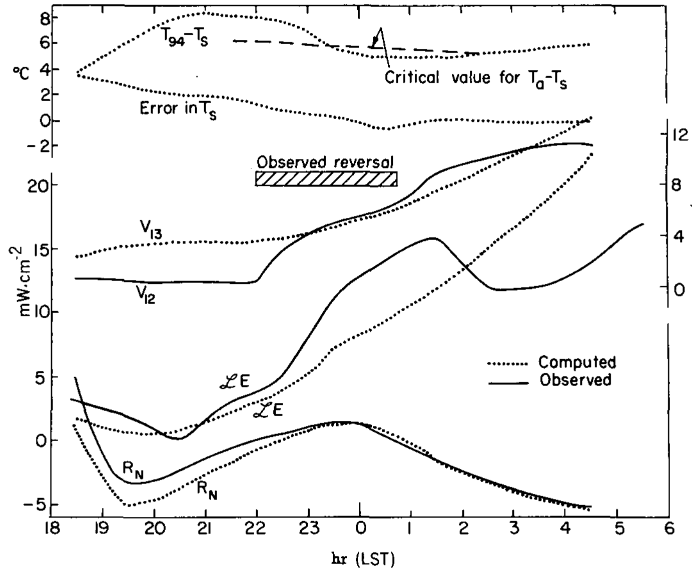

FIGURE 1.—Comparison of observed values (solid lines) and computed values (dotted lines) of net radiation, R_N, evaporation, $\mathcal{L}E$, and surface wind speed, V. The suffix is the height(m) to which V refers. The computed temperature difference between 94 m and the surface is compared to the critical value of T_a-T_s (dashed line). The time of the observed radiation reversal is shown by the hatched area.

We thus attempted to simulate events at Aspendale during the night of Oct. 4–5, 1970, taking for granted the working of large-scale dynamics, in regard to surface pressure gradient, advection, and downward radiation, a function of the cloud amount and temperature. No variation of geostrophic wind with height was permitted.

4. RESULTS

The focus is on the hours of darkness, and only the data, observed and computed, from 1830 to 0430 LST are presented. Strict verification is only possible for evaporation and net radiation (the downward component of which was taken as given, so that the ability to predict surface temperature is being tested), but the inequality [eq (3)] given in Clarke (1972) can be checked if T_c is taken to be the air temperature computed at 3000 m, the observed height of the cloud.

Figure 1 shows the course of latent heat flux, net radiation, and surface wind, both measured and computed; the error in computed surface temperature; and $T_{94}-T_s$, the temperature increment from the surface to 94 m, the height of the third model level.

It is seen that the surface inversion (over the lowest 94 m) meets the requirement [eq (3) of Clarke 1972] over the first part of the radiation reversal only; but if the inversion is measured to 174 m, the next model level, this requirement is satisfied over the whole 3 hr of reversal.

As a check on the computed temperature and humidity profiles, these were used to assess R at 2330 LST with the insertion of clouds at 3000 m by means of an Elsasser diagram. Agreement was within 1 percent of the measured value of 43.7 $mW \cdot cm^{-2}$.

Verification of the upper wind profile against that measured at 0300 LST on the 5th at Laverton is not considered to be informative because of the lack of thermal wind in the model and the presence of a rather large thermal wind from the west indicated by the data. Therefore, although the speed profile is well predicted, that of direction is not.

5. CONCLUSIONS

The modeling appears to be fairly successful. The radiation reversal, depending on the prediction of surface temperature, is reproduced, and evaporation is within reach of the measured values, which are considered reliable within 0.001 in./hr (≈ 1.7 $mW \cdot cm^{-2}$ in terms of energy flux). Equation (1) of Clarke (1972) is of course satisfied, since this is built into the model, and his inequality (3) is reasonably satisfied, considering the approximations in the radiation model.

The poorest agreement is in surface temperature and wind during the period of calm observed at Aspendale before 2200 LST. This was not reproduced by the model, but neither was it found in the records of three other land stations within 40 km. These measured wind speeds were closely similar to those computed. It is probably beyond our present capability to model such a local calm period with moderate geostrophic wind.

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